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Key Points:

- Past Pacific plate motions constrain asthenospheric viscosity
- Hot spot drift and Pacific bend event duration do not hamper inferences
- Asthenosphere response to shear over kyr is consistent with that over Myr

Supporting Information:

- Readme
- Supplementary material

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Pacific plate-motion change at the time of the Hawaiian-Emperor bend constrains the viscosity of Earth's asthenosphere

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Abstract Important constraints on asthenospheric viscosity come primarily from modeling the glacial rebound of the past 20 kyr, but remain somewhat loose because of the intrinsic resolving power of these models. We obtain narrower bounds by building on the notion that the asthenosphere also controls the ability to change plate motions over Myr. We focus on the Pacific kinematic change at the time of the Hawaiian-Emperor bend event, which is linked to the coeval inception of subduction in the Western Pacific. We sample plausible asthenospheric viscosity and thickness values by requiring the rate at which torque varied to generate the observed kinematics consistent with the nature of subduction initiation. Uncertainties on the bend event duration and the occurrence of Pacific hot spots drift do not hamper our results that suggest that the asthenosphere viscous response to vertical shear over kyr is consistent with that to horizontal shear over Myr.

1. Introduction

Barrell, in his seminal work on gravity anomalies and isostatic compensation, was the first to refer to the presence of—in his own words—a sublithospheric *sphere of weakness*, or *asthenosphere*, that could accommodate vertical stress variations due to erosion and sedimentation [*Barrell*, 1914]. The advent of the theory of plate tectonics in the 1960s [e.g., *Morgan*, 1968] revamped the discussion on the asthenosphere and its viscosity because the history of plate-motion changes inferred from the ocean-floor magnetization suggested the presence of a mechanically weak layer on top of which plates glide. This inference was in line with the first observations of attenuation of seismic waves through the upper mantle [*Oliver and Isacks*, 1967]. Several authors speculated later that Earth's asthenosphere owes its origin to the presence of melt and water at ranges of temperature and pressure associated with sublithospheric depths [e.g., *Bai and Kohlstedt*, 1992; *Karato*, 2012].

Since the original work of *Haskell* [1935, 1937], constraints on the radial viscosity structure of the mantle, and thus also on the asthenospheric viscosity, come primarily from observations of glacial rebound [see *Mitrovica et al.*, 2007, for a review]. They inform us about the rates at which regions such as Fennoscandia, Canada, and Antarctica rise in response to melting of thousand-kilometer-wide ice sheets since the last glacial maximum, some 10 kyr ago. Glacial rebound models classically employ a seismologically defined layering of Earth's mantle: a low-viscosity zone above the 670 km seismic discontinuity, and a more viscous lower mantle extending from there to the core-mantle boundary. Further subdivisions of the low-viscosity zone have also been successfully explored [e.g., *Lambeck et al.*, 1996; *Lambeck and Johnston*, 1998]. In revisiting the work of *Haskell* [1935], *Mitrovica* [1996] made the important point that the upper region of Earth's mantle may feature significant increase in viscosity with depth, yet with the average viscosity being consistent with the Haskell value of 10^{21} Pa s. Such an inference implied that the spectrum of plausible radial viscosity profiles may be wider than expected. Results from numerous studies on glacial rebound modeling [e.g., *Lambeck et al.*, 1990; *Fjeldskaar and Cathles*, 1991] favor a strong asthenosphere/mantle viscosity contrast as large as 3 orders of magnitude. This inference is also supported by inversions performed jointly with mantle circulation models [*Mitrovica and Forte*, 2004] and by independent results obtained from modeling long–wavelength anomalies of the geoid [e.g., *Ricard et al.*, 1984].

In reviewing results from previous studies, *Paulson and Richards* [2009] pointed out that despite the large amount of data gathered, from the original work of *Haskell* [1935] to the GRACE mission, the sensitivity of glacial rebound models to the radial viscosity structure of Earth's mantle is such that significantly-different solutions are equally warranted by the observations. They showed that rebound modeling ultimately constrains the viscosity contrast between asthenosphere and lower part of the upper mantle to be proportional to the cube of the asthenospheric thickness. That is, a layered upper mantle model featuring a low-viscosity asthenosphere would yield a similar fit to the rebound data as a uniform upper mantle featuring a viscosity roughly equal to the depth-integrated value of the layered model. One conclusion that can be drawn from these inferences is that while glacial rebound modeling is key in illuminating the mechanical stratification of Earth's mantle, the associated resolving power with respect to the asthenospheric viscosity is perhaps not as high as one may wish. In a notable study, *Gaboret et al.* [2003] explored the impact that the inclusion of a low-viscosity ($\sim 10^{19}$ Pa s) layer in models of present-day mantle convection would have on predictions of seismic anisotropy beneath the Pacific basin. They found an agreement with observed azimuthal directions within $\sim 50°$. While this advance is encouraging, the issue of resolving asthenospheric viscosity and thickness still persists today.

Earth's asthenosphere plays an important role not only in regulating the viscous response to vertical shear over kyr time scales associated with glacial rebound but also in modulating the dynamics of the coupled plates/mantle system, which involve predominantly horizontal shear over Myr periods. The asthenospheric viscosity impacts on the circulation planform of mantle convection [e.g., *Bunge et al.*, 1996], on the pattern of plume hot spots activity recorded on the aging ocean floor [*Boschi et al.*, 2007]; and it is thought to influence the onset of plate tectonics on terrestrial planets [e.g., *O'Neill et al.*, 2007]. On a more fundamental level, the asthenosphere impacts on the ability of tectonic forces to change plate motions. As plates glide over the low-viscosity asthenosphere, their base undergoes shear tractions that contribute to the plate torque balance. The torque associated with these tractions depends on the plate's angular velocity, its basal area, and the asthenospheric viscosity. The dynamic balance dictates that such a torque be equal and opposite to the net of all other torques acting upon a plate—an inference that clarifies the dominant role of the asthenosphere in the dynamics of plates. By the same logic, plate motion changes reconstructed from observations of the ocean-floor magnetization may be used to probe the mechanical characteristics of the asthenosphere.

Here we build on such a notion to further narrow the constraints on Earth's asthenosphere available from glacial rebound modeling; but we remain aware of the different dynamics (vertical shear over kyr versus horizontal shear over Myr) associated with these geological processes. We focus on the well-documented change of Pacific plate motion at the time of the Hawaiian-Emperor bend (HEB) [e.g., *Wessel and Kroenke*, 2008], which resulted from the inception of the Izu-Bonin, Mariana, and Tonga-Kermadec subductions in the Western Pacific around the same time [*Faccenna et al.*, 2012]. First, we consider the rate at which torque needed to be built upon the Pacific plate in order to produce the kinematic change reconstructed by *Wessel and Kroenke* [2008]. We sample the magnitude of such a *torque variation-rate* for broad ranges of asthenospheric viscosity and thickness values. Second, we compare these to estimates of the torque variation-rate available from subduction initiation along the Western Pacific trenches at the HEB time. This allows us to identify plausible ranges of asthenospheric viscosity and thickness that are narrower than, but still consistent with, what is inferred from glacial rebound observations. We then take into account the possibility that the reconstructed Pacific kinematic change is to some degree biased by our lack of detailed knowledge of the Pacific hot spots drift around the time of HEB. Finally, we discuss the impact of these inferences on the notion that the viscous response of Earth's asthenosphere over time scales of kyr is consistent with that over Myr.

2. Kinematic Constraints

As seafloor volcanos originate from upwelling in the Earth's mantle [*Wilson*, 1963], the most compelling evidence that the Pacific plate motion changed significantly during the mid-Cenozoic [e.g., *Wessel and Kroenke*, 2008] are (i) the striking change of direction of the Hawaiian-Emperor volcanic chains, visible today on the Pacific ocean floor [e.g., *Smith and Sandwell*, 1997] together with (ii) the inference from isotopic dating techniques that these volcanos become older as one moves away from the present-day hot spot beneath Hawaii [e.g., *Clague and Dalrymple*, 1987; *Sharp and Clague*, 2006]. *Tarduno et al.* [2003] proposed that the Hawaiian-Emperor volcanic track in reality recorded chiefly the drift of the Hawaiian plume—in response to North Pacific ridge capture and release during Early Cenozoic—and perhaps to a minor degree the motion change of the Pacific plate. Nonetheless, the bend event is also recorded on the track of the Louisville volcanic chain and the kinematics of the Pacific plate at the HEB time remains today the most notable example of plate-motion change in the geoscience literature.

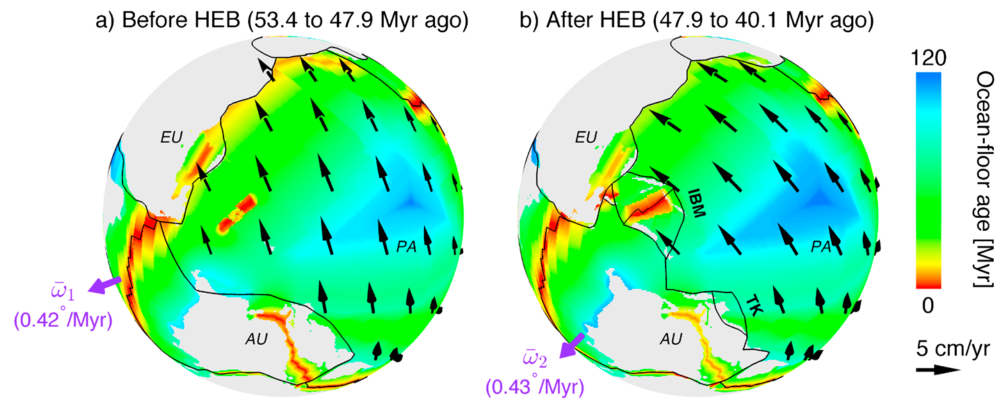

Figure 1. Pacific plate (PA) absolute motions [*Wessel and Kroenke*, 2008] (a) before and (b) after the Hawaiian-Emperor bend (HEB) event. In purple are Euler vectors; black arrows are the associated surface velocities. Plate boundaries are in black. In color is the ocean-floor age. AU = Australian plate; EU = Eurasian plate. Visible features are (i) the inception of the Izu-Bonin and Mariana (IBM) and Tonga-Kermadec (TK) subduction systems, (ii) the transition from transform to oblique convergent regime in the Western Pacific basin, and (iii) the change of direction of PA motion.

Recent improvements in dating techniques based on the $^{40}Ar/^{39}Ar$ isotopic ratio led *Sharp and Clague* [2006] to revise the time and duration of the bend phase, which they estimated to have occurred over a period $\Delta t_B = 1.2 \pm 0.3$ Myr around 47.3 Myr ago, rather than 43 Myr ago—as previously inferred [e.g., *Clague and Dalrymple*, 1987]. *Wessel et al.* [2006] pointed out the synchroneity of the HEB with several other tectonic events around the Pacific Rim, such as triple-junction reorganization in the South Pacific, changes in the Kula/Pacific spreading rates in the north, and bent fracture zones along the Pacific/Farallon margin to the east. Further, *Whittaker et al.* [2007] found evidence of a major Australia/Antarctica reorganization at the time of the HEB. These observations would suggest a predominantly tectonic origin of the Pacific kinematic change.

Figure 1 shows the Pacific plate motion change at the time of the HEB, as reconstructed by *Wessel and Kroenke* [2008] for stages 53.4 to 47.9 Myr ago (before the HEB—Figure 1a) and 47.9 to 40.1 Myr ago (after the HEB—Figure 1b). In purple are stage Euler vectors for the Pacific rigid motion. We obtained these by differentiating the finite rotations of *Wessel and Kroenke* [2008] with respect to time, according to the algebra of rigid rotation matrices. Black arrows are the associated surface velocities, which show a significant change in direction, from northward to northwestward. This is perhaps the most rapid kinematic change involving a major plate over the past ~150 Myr ever documented [e.g., *Müller et al.*, 2008]. In color is the age of the ocean floor at the middle of the associated stage, as reconstructed by *Müller et al.* [2008] from a global compilation of magnetic isochrons. Black lines are past plate margins from the compilation of continuously closing plates of *Gurnis et al.* [2012]. Visible features are the switch from transform to oblique convergent regime in the Western Pacific, the inception of the Izu-Bonin, Mariana, and Tonga-Kermadec systems and the consequent subduction of relatively old (~60 Myr), and thus thick (~100 km), ocean floor. The striking coincidence of the HEB kinematic change of the Pacific plate with the inception of subduction systems in the west led *Faccenna et al.* [2012] to demonstrate the geodynamical plausibility of these events being in fact linked. The initiation of subduction along the Izu-Bonin, Mariana, and Tonga-Kermadec trenches provided the additional torque upon the Pacific plate that was necessary to divert its motion from northward to northwestward [*Faccenna et al.*, 2012].

3. Methods and Results

The Pacific kinematic change at the time of the HEB provides the opportunity to probe viscosity and thickness of Earth's asthenosphere. We build on the previous work of *Iaffaldano et al.* [2012, 2013] who provided a method that makes use of kinematic reconstructions to estimate the rate at which torque needs to be built upon tectonic plates—also referred to as *torque variation-rate*—in order to explain their kinematic changes. Such a method bears upon the well-known torque balance of rigid bodies rotating on top of a fluid, highly viscous sphere; and consists of differentiating the torque balance equation at two points in geological time. To gain geodynamical insight, these estimates may then be compared with the rates at which geological processes evolve and build torque upon tectonic plates.

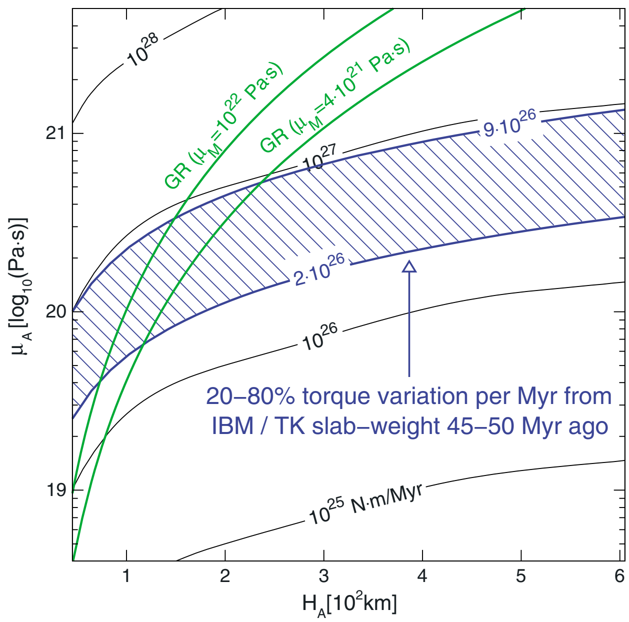

Figure 2. In black are contours of values of torque variation-rate necessary upon PA to explain its kinematic change at the HEB time, sampled for ranges of asthenospheric thickness (H_A) and viscosity (μ_A). Thick blue contours, instead, correspond to rates upon PA associated with inception of Western Pacific subductions, assuming that 20% to 80% of the slab weight is transmitted to PA. Blue-striped area represents ranges of asthenospheric viscosity and thickness for which the torque variation-rate required upon PA to explain its kinematic change falls within the range of what the inception of the IBM and TK systems may provide. Green lines show sets of parameters (μ_A, H_A) consistent with modeling observations of glacial rebound (GR). Right-hand side profile corresponds to assuming the viscosity of the lower part of the upper mantle (μ_M) to be $4 \cdot 10^{21}$ Pa s; left-hand side profile corresponds to $\mu_M = 10^{22}$ Pa s.

In the specific case of the Pacific plate, the Euler vectors in Figure 1 reflect the readjustment of torques acting upon the plate during the HEB event, following the inception of the Izu-Bonin, Mariana, and Tonga-Kermadec systems. As *Wessel and Kroenke* [2008] reconstructed the Pacific motion in an absolute reference frame fixed with the deep mantle, these Euler vectors allow the estimation of the torque generated upon the plate by its shearing over the asthenosphere [e.g., *Iaffaldano et al.*, 2012, 2013]. Such a torque depends on the asthenospheric viscosity and also on the thickness of this layer, since the flow induced by the Pacific plate motion within the low-viscosity channel is capable of shearing the latter down to its boundary with the more viscous part of the upper mantle. It is from such a notion that stems the geodynamical definition of asthenosphere as the sublithospheric, low-viscosity region of the mantle where surface plate motions induce significant flow of Couette type. By virtue of the dynamic balance, the torque associated with plate shearing over the asthenosphere must be equal and opposite to the net of coeval torques generated upon the Pacific plate by other forces, such as the net pull of the subducted Pacific lithosphere or the frictional resistance along the brittle plate margins, among others. In other words, the shear torques associated with $\bar{\omega}_1$ and $\bar{\omega}_2$ of Figure 1 must balance the net of all other torques upon the Pacific plate, respectively before and after the HEB. The difference of these shear torques must equal the additional torque variation built upon the plate during the HEB, so that its motion evolves through time as reconstructed by *Wessel and Kroenke* [2008]. Further, such a torque variation occurs over a period of time equal at most to the duration of the HEB event, inferred to be $\Delta t_B = 1.2 \pm 0.3$ from independent observations [*Sharp and Clague*, 2006]. All these inferences allow us to sample the magnitude of the torque variation-rate required upon the Pacific plate during the HEB (see supporting information), as a function of assumed asthenospheric viscosity (μ_A) and thickness (H_A) (Figure 2, black contour lines). The torque variation rate increases for greater values of μ_A, because greater shear tractions at Pacific plate base must be overcome in order to change its kinematics. Instead, greater values of H_A yield smaller strain rates within the asthenosphere and therefore smaller resistance to changing the Pacific plate motion. Consequently, torque variation rates decrease for greater values of H_A.

We compare these estimates with the rate at which subduction initiation along the Izu-Bonin, Mariana, and Tonga-Kermadec systems may build torque upon the Pacific plate. As prior to this tectonic event no slabs are attached to the Pacific plate along these trenches, such a torque variation-rate arises from the net weight—that is, the volume integral of the negative buoyancy—of the newly subducted Pacific lithosphere within a unit time. One should in theory reduce the net weight by the viscous force exerted upon the slab as it plunges into Earth's asthenosphere. However, this is demonstrably 1 to 2 orders of magnitude smaller than that from the slab negative buoyancy [e.g., *Davies*, 1999], and thus can be comfortably neglected. To compute the slab net weight per unit time, we implement the well-known *slablets* method outlined by *Ricard et* *al.* [1993]. It consists of integrating the slab negative buoyancy—set here to 70 kg/m^3 [e.g., *Faccenna et al.*, 2012]—over the slab volume increase within a unit time (see supporting information). We use 1 Myr intervals, which is short enough to reasonably neglect slab buoyancy changes due to heat diffusion. At any point along the Izu-Bonin, Mariana, and Tonga-Kermadec systems, we compute the volume of slab subducted in 1 Myr from the age of the oceanic lithosphere [*Müller et al.*, 2008]—through the half-space cooling model [e.g., *Davies*, 1999]—and the surface velocity at that particular point along the trench systems [*Wessel and Kroenke*, 2008]. In line with previous studies [e.g., *Conrad and Lithgow-Bertelloni*, 2002; *Faccenna et al.*, 2012], we assume that the net weight pulls the trailing plate in a direction normal to the local trench strike. Finally, we compute the torque variation-rate upon the Pacific plate associated with the net weight of the newly subducted slab over 1 Myr. We systematically repeat this procedure using tectonic information about the Pacific plate (i.e., age, velocity, and shape of plate margins) from 50 to 45 Myr ago, which is a conservative estimate of the duration of the HEB phase (i.e., roughly twice the estimate by *Sharp and Clague* [2006]), and elect to take the maximum torque variation-rate within this period as most representative. We note that the age of the subducting slab is known with relatively high confidence from magnetic reconstructions. Thus, slab buoyancy and thickness are also known with confidence, as they are linked to the slab age through the half-space cooling model. The main uncertainty in estimating the net weight of subducting slabs per unit time comes from the uncertainty on plate velocities at the trench. In the cases of the Izu-Bonin, Mariana, and Tonga-Kermadec systems, we estimated an uncertainty of 10% from the covariances associated with the kinematic reconstruction of *Wessel and Kroenke* [2008]. The additional torque available to the Pacific plate per Myr around the time of the HEB may be at most equal to, but it is very likely less than, the torque associated with the net weight of subducting slabs in the Western Pacific. In fact, it is generally unclear how much of the net weight of slabs is consumed in the process of bending the trailing plate, with the remaining part instead contributing chiefly to a pull. Previous estimates [e.g., *Conrad et al.*, 2004; *Buffett*, 2006; *Faccenna et al.*, 2012] yield an admittedly wide range of 30 to 70%. In order to take this into account, as well as the uncertainty on past plate motions, we assume a wider range of 20 to 80% to estimate the lower and upper limits of the torque variation-rate that the Izu-Bonin, Mariana, and Tonga-Kermadec systems may possibly contribute altogether to the dynamics of the Pacific plate. These are $2 \cdot 10^{26}$ and $9 \cdot 10^{26}$ N m/Myr respectively. Thick blue lines in Figure 2 are the contours corresponding to these values. The blue-striped area in between shows combinations of parameters (μ_A, H_A) for which the torque variation-rate required to explain the Pacific kinematic change falls within the range of what the inception of the Western Pacific subduction systems may provide upon the plate. That is, they are geodynamically plausible combinations of parameters. In contrast, asthenospheric viscosity and thickness anywhere above or below the striped area imply that the torque variation-rate required upon the Pacific plate is, respectively, greater or smaller than what the Izu-Bonin, Mariana, and Tonga-Kermadec systems may provide. We deem these parameter combinations as geodynamically implausible.

4. Discussion

These inferences provide the ability to constrain the asthenospheric viscosity within some 2 orders of magnitude, regardless of its thickness. In fact, the asthenospheric viscosity must be in the range $\sim \cdot 10^{19}$ to $\sim 10^{21}$ Pa s in order for the inception of the Western Pacific subductions to efficiently redirect the Pacific plate motion at the time of the HEB event (Figure 2). We compare these values to those inferred from glacial rebound modeling; although we remain aware of the main difference between the dynamics associated with these geological processes: that is, glacial rebound involves a viscous response of Earth's asthenosphere to vertical shear over time scales of kyr, while plate-motion changes determine horizontal shear within the asthenosphere over time scales of Myr. Green profiles in Figure 2 show the key finding of numerous studies of the observed glacial rebound [see *Paulson and Richards*, 2009, and references therein for a review]: that rebound modeling constrains the viscosity contrast between asthenosphere and lower part of the upper mantle to be proportional to the cube of the asthenosphere thickness. The green profile on the left-hand side shows sets of parameters (μ_A, H_A) warranted by glacial rebound observations when the viscosity of the lower part of the upper mantle (μ_M) is assumed to be 10^{22} Pa s; while the right-hand side profile corresponds to $\mu_M = 4 \cdot 10^{21}$ Pa s. We require that both observational constraints—glacial rebound on the one side, Pacific plate kinematics on the other—be satisfied. This constrains the asthenospheric viscosity between $4 \cdot 10^{19}$ (intersection of left-hand side green profile with lower blue contour) and $5 \cdot 10^{20}$ Pa s (intersection of right-hand side green profile with upper blue contour). Such a range is narrower than is inferred

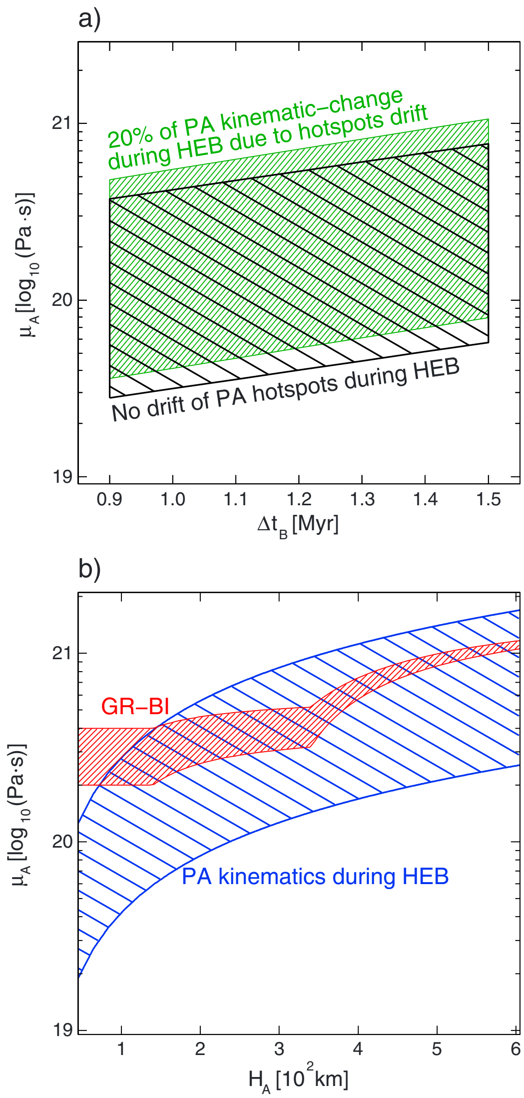

Figure 3. (a) In black: range of asthenospheric viscosity (μ_A), sampled for durations of the bend event (Δt_B), that satisfies constraints from both glacial rebound (GR) observations and PA kinematic change at the HEB time. In green: same as black, resampled assuming that 20% of the PA kinematic change owes to drift between the Hawaiian and Louisville hot spots. (b) In blue: required range of μ_A in order for the inception of the IBM and TK systems to drive the PA kinematic change at the HEB time, as a function of H_A. In red: depth-averaged viscosity, from the base of the lithosphere to the depth associated with H_A, obtained from the glacial rebound model of *Lambeck et al.* [1996], which is constrained with the high-quality rebound record of the British Isles (GR-BI).

from glacial rebound [*Paulson and Richards*, 2009] or plate kinematic modeling alone. Glacial rebound data constrain the average upper mantle viscosity beneath continents to be a factor 3–5 larger than beneath oceanic basins [*Lambeck and Chappell*, 2001]. However, we note that, to very first order, such a factor disappears in taking the ratio between viscosities of asthenosphere and lower part of the upper mantle. This may make the inference of *Paulson and Richards* [2009] applicable also to the mantle beneath oceanic lithosphere. Further, we note that the extent of the inferred range is in line with the independent study by *Mitrovica and Forte* [2004], but we emphasize that fixity of the low-viscosity layer depth is not enforced in our inference. In fact, from Figure 2 one also derives that the asthenosphere may be up to 250 km thick.

Because our inferences are based on comparing rates at which torques upon the Pacific plate vary through time, they rely significantly on the duration of the HEB event $\Delta t_B = 1.2 \pm 0.3$ Myr [*Sharp and Clague*, 2006]. We explore how sensitive they are to the uncertainty associated with Δt_B: we vary the assumed duration of the HEB phase within the range proposed by *Sharp and Clague* [2006] and resample plausible asthenospheric viscosities that jointly satisfy glacial rebound and Pacific kinematics constraints. We show these in Figure 3a (black-striped area) and conclude that the range of plausible viscosity values does not vary significantly within the uncertainty of Δt_B. A number of observations support a tectonic origin for the mid-Cenozoic Pacific kinematic change (see section 2), something on which we base our inferences on the asthenospheric viscosity. While testing the impact of other tectonic processes on the Pacific plate motion at the time of the HEB is beyond the main focus of this study, we note that subduction is among the key geological processes controlling plate motions and their temporal changes. On this basis, one could argue that other tectonic processes at work at the time of the HEB would provide less torque variation-rate, and thus be of second-order importance than—or even overwhelmed by—subduction initiation.

Two outstanding questions remain unanswered to date: (i) how much did the Hawaiian and Louisville plume hot spots drift around the time of the HEB? (ii) Assuming either or both plumes did drift, as suggested by *Tarduno et al.* [2003], how would this knowledge impact reconstructions of the absolute Pacific motion [e.g., *Wessel and Kroenke*, 2008]? While more radioisotopic and paleomagnetic data—particularly from the Louisville hot spot—will be needed to address such questions in the future, here we assume that hot spots drift may impact on the reconstructed Pacific kinematic change by 20% at most. Such a number is not arbitrarily chosen; rather it takes into account recent results from geodynamic modeling of plume drift within a convecting mantle. Specifically, *Davies and Davies* [2009] found that mantle plumes may drift at rates of ∼1 cm/yr, which corresponds to some 20% of the Pacific kinematic change reconstructed by *Wessel and Kroenke* [2008] around the HEB event. This is also in line with results from earlier studies that inferred, from the observed curvature of the ocean floor volcanic track, a relatively small amount of drift of the Hawaiian plume [*Griffiths and Richards*, 1989]. Accordingly, we recompute plausible ranges of asthenospheric viscosity after reducing by 20% the kinematic change described by Euler vectors $\bar{\omega}_1$ and $\bar{\omega}_2$. Results (Figure 3a, green-striped area) are remarkably similar to those obtained assuming no hot spots drift (Figure 3a, black-striped area). This indicates that the limited knowledge on the Cenozoic drift of plume hot spots in the Pacific in fact does not impact significantly on geodynamical inferences of the asthenospheric viscosity.

We required that both observational constraints from glacial rebound and Pacific plate kinematics be satisfied in order to derive geodynamically plausible ranges of asthenospheric viscosity and thickness. In doing so, we made the following assumption: that the viscous response of the asthenosphere to vertical shear associated with glacial rebound on the one hand, and horizontal shear associated to plate-motion changes on the other hand, remains the same over the two significantly different time scales: the first of kyr and the second of Myr. Such an assumption is at the core of glacial rebound modeling [e.g., *Cathles*, 1975], but to date it is yet to be verified. Before we use our inferences on the Pacific plate dynamics to do so, we recall that here a geodynamical definition of asthenosphere thickness has been adopted; and that this in general differs from the seismologically-defined layering of Earth's mantle typically employed in glacial rebound modeling. We elected to define as asthenosphere the low-viscosity region comprised between the lithosphere base and the depth at which the Couette-type viscous flow induced by surface plate motions decays to zero, due to mechanical contrast with the more viscous upper part of Earth's mantle. This depth is a free parameter in our models, as we are aware of evidence from seismic observations in support of the notion that the thickness of the asthenosphere beneath oceans might differ from the one underneath continents [e.g., *Gung et al.*, 2003]. In glacial rebound modeling, the asthenosphere is also defined as a low-viscosity layer extending from the base of the lithosphere to some depth above the seismically observed 670 km discontinuity [*Lambeck et al.*, 1996]; but layering of Earth's mantle is kept fixed in these models, and only the associated viscosities are free parameters of the problem. We aim at testing the consistency of the asthenospheric viscosity value over different time scales regardless of how the asthenosphere is defined. In Figure 3b we report in blue the inferred range of asthenospheric viscosity (μ_A) required in order for the inception of the Western Pacific subductions to drive the Pacific kinematic change at the time of the HEB, as a function of the asthenospheric thickness H_A—according to the geodynamical definition. For each thickness value, the range of plausible viscosities accounts for our uncertainty on (i) how much of the net slab weight contributes to pulling the trailing plate (see Figure 2), (ii) the reconstructed Pacific plate motion at the time of the HEB, and (iii) the duration of the bend event (see Figure 3a). Further, we recall that the viscosity estimates so derived are independent of rebound analyses. We compare these values with the five-layer regional model for radial viscosity stratification of *Lambeck et al.* [1996], inferred from the high-quality rebound record of the British Isles (Figure 3b, in red). For each value of the geodynamically defined thickness H_A, we report in red the viscosity range warranted by the model of *Lambeck et al.* [1996]. This range consists of the average viscosity, between the base of the lithosphere and the base of the geodynamically-defined asthenosphere. This time around we note that the viscosity range so derived is independent of plate-motion reconstructions. Yet it is consistent to large degree with the range inferred from the Pacific plate dynamics, regardless of the depth at which one may elect to place the geodynamical asthenosphere base (Figure 3b). If the somewhat lower asthenospheric viscosity estimate obtained from isostatic rebound modeling of continental and ocean-island sites far from the formerly glaciated regions are adopted [*Lambeck et al.*, 2002], then this agreement is even more consistent. This suggests that the asthenosphere viscous response to vertical shear over kyr is consistent with that to horizontal shear over Myr.

5. Conclusions

Geodynamical inferences derived from the kinematic change undergone by the Pacific plate at the time of the Hawaiian-Emperor bend event, combined with previous results from modeling observations of glacial rebound, constrain the viscosity of Earth's asthenosphere in range $\sim 4 \cdot 10^{19}$ to $\sim 5 \cdot 10^{20}$ Pa s. Uncertainties on the duration of the bend phase—which could have lasted 0.9 to 1.5 Myr around 47.3 Myr ago—and the drift of the Hawaiian and Louisville hot spots—which could potentially account for up to 20% of the Pacific kinematic change at that time—do not impact significantly on the inferred values of asthenospheric viscosity. Results suggest that the asthenosphere viscous response to vertical shear over time scales of thousand years is consistent with that to horizontal shear over million years.

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